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(Article begins on next page)

1 **PRESSURE-TEMPERATURE-TIME-DEFORMATION PATH OF KYANITE-BEARING**
2 **MIGMATITIC PARAGNEISS IN THE KALI GANDAKI VALLEY (CENTRAL NEPAL):**
3 **INVESTIGATION OF LATE EOCENE-EARLY OLIGOCENE MELTING PROCESS**

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37 pseudosection modelling, Zr-in-rutile thermometry

38 **1. Introduction**

39 The Himalaya-Tibet orogenic system is often regarded as the classic “type” of a continental
40 collisional belt. This belt is characterized by a continuity of both main tectonic units and
41 discontinuities for almost 2400 km along strike. Among these units, the Greater Himalayan

42 Sequence (GHS) represents the now-exhumed metamorphic core of the Himalayan orogenic
43 pile (Hodges, 2000), where medium- to high-grade metamorphic rocks and migmatites occur
44 (*e.g.* Hodges, 2000). The pressure (P) – temperature (T) – time (t) paths of migmatites can
45 carry valuable information on the thermal history and rheological evolution of collisional
46 belts (*e.g.* Searle, 2013; Yakymchuk and Brown, 2013; Hallett and Spear, 2014; Weinberg and
47 Hasalová, 2015). Melting has profound effects on the rheological properties of rocks such as a
48 dramatic reduction of their viscosity even if only small volumes of melt (> 5%) are present
49 (Rosenberg and Handy, 2005; Jamieson et al., 2011). This “melt-weakening effect”, which
50 should have affected a large part of Himalayan high-grade rocks, together with focused
51 denudation/erosion, is the main prerequisite to apply channel flow models (Beaumont et al.,
52 2001; Jamieson et al., 2004; Jamieson et al., 2011). These models can explain the evolution
53 and exhumation of high-grade metamorphic rocks, although the mechanisms, the productivity
54 and the timing of melting are far from being well understood in the Himalaya despite
55 numerous experimental and empirical investigations made by many authors (*e.g.* Patiño-
56 Douce and Harris, 1998; Harris et al., 2004; Guilmette et al., 2011; King et al., 2011; Searle
57 2013; Palin et al., 2014; Imayama et al., 2012; Groppo et al., 2010, 2012 and references
58 therein). Several melting stages have been proposed for this mountain belt occurring at
59 different times. For example, in the Sikkim region (Fig. 1) Rubatto et al. (2013), using
60 monazite and zircon petrochronometers, identified two portions of the migmatitic GHS, which
61 experienced partial melting and cooling at different times: structurally lower migmatites
62 experienced melting and peak metamorphic conditions at c. 31-27 Ma, earlier than the
63 structurally higher ones (26-23 Ma). Also Kohn et al. (2005) and Imayama et al. (2012)
64 identified a clear diachronic evolution of GHS melting in the Langtang section, (Central Nepal)
65 and in the Tamor-Ghunsa transect (Eastern Nepal) respectively (Fig. 1). In both cases the

66 structurally higher samples experienced melting earlier than the lower ones (20 Ma vs. 15 Ma
67 in Kohn et al., 2005, and 33-28 Ma vs. 21-18 Ma in Imayama et al., 2012).

68 Here, we focus on kyanite-bearing migmatitic paragneiss in the Kali Gandaki valley (Central
69 Himalaya, Fig. 1) located in a key structural position (Vannay and Hodges, 1996) due to its
70 closeness to the Main Central Thrust (MCT). Recently, Carosi et al. (2014a) identified within
71 garnets coming from the same outcrop, the presence of crystallized melt inclusions, referred
72 as “nanogranites” (Carosi et al., 2014a and references therein) showing a peculiar chemical
73 composition of high-Ca melts (tonalites). According to Patiño-Douce and Harris (1998) and
74 Prince et al. (2001) this melt types is interpreted as result of high-pressure (HP) melting,
75 possibly in the presence of free water and through K-feldspar-absent reactions (see also King
76 et al., 2011; Palin et al., 2014; Weinberg and Hasalová, 2015).

77 We present pressure (P) - temperature (T) - deformation (D) path for the kyanite-bearing
78 migmatitic paragneiss, which is based on the pseudosection approach (*e.g.* Vance and Mahar,
79 1998) and trace element thermometry (*e.g.* Spear and Pyle, 2002; Hallett and Spear, 2014)
80 coupled with careful meso- and micro-structural observations. Texturally and chemically
81 controlled *in situ* U-Th-Pb monazite ages add important time constraints to this path (*i.e.* P-T-
82 t-D paths). It has already turned out that *in situ* U-Th-Pb monazite dating, with the advantage
83 that textural and chemical “grain-fingerprints” are preserved, coupled with the pseudosection
84 approach is a powerful tool to temporally bracketing the P-T-D history of metamorphic rocks
85 (see, *e.g.*, Williams and Jercinovic, 2002, 2012). At last, possible tectonic implications will be
86 discussed.

87 **2. Geological overview of the Himalayan Belt**

88 The Himalayan mountain belt (Fig. 1a) is the result of the collision between the Asian and
89 Indian continental plates around 55-50 Ma (Hodges, 2000) after the break-up of Gondwana

90 and a long last-standing Andean-type active margin, which resulted from subduction of Neo-
91 Tethys oceanic crust below the Lhasa Block, the intrusion of large (mainly I type) granitoid
92 bodies and accretion of arc terranes. The precise age of the India-Asia collision as well as a
93 possible diachronic collision is still under debate in the geological literature (*e.g.* Najman et al.,
94 2010 and references therein).

95 Since the collision occurred, crustal rocks representing the northern front of the Indian plate
96 have experienced a complex deformative and metamorphic history, building up part of the
97 Himalaya as we observe this mountain range nowadays (Gansser, 1964; Heim and Gansser,
98 1939; Hodges, 2000; Yin, 2006).

99 The structural architecture of the Himalayan chain is made by different tectono-metamorphic
100 units (Fig. 1a) separated by important tectonic structures (regional scale reverse and normal
101 shear zones). In an “ideal” profile from south to the north, the following tectonic units are
102 present according to Hodges (2000):

103 1) The Siwalik Unit (SU) is made of recent sediments tectonically sandwiched between the
104 undeformed molasse of the Ganga plain and the upper unit, the Lesser Himalayan Sequence
105 (LHS). The lower tectonic contact is a top-to-the south thrust system referred as Main Frontal
106 Thrust (MFT), whereas the Main Boundary Thrust (MBT) separates the SU from the LHS.

107 2) The LHS is made of Lower Proterozoic to early Palaeozoic low to medium grade
108 metasediments and meta-igneous rocks. This unit is tectonically overlain by medium- to high-
109 grade metamorphic rocks of the GHS *via* a regional top to the S-SW (thrust sense) shear zone,
110 called Main Central Thrust (MCT, Figure 1). Since the MCT is not a single thrust surface, but a
111 thick ductile to brittle shear zone with a variable thickness (100 m up to several km, Searle et
112 al., 2008), we prefer the term Main Central Thrust Zone (MCTZ) to identify its sheared rocks.

113 According to Stephenson et al. (2001) both GHS and LHS rocks are ductily sheared by the MCT

114 activity, with the latter rocks ductily incorporated in the MCTZ during the shear zone activity
115 (widening of the shear zone towards the S).

116 3) The GHS consists of Late Proterozoic to Cambrian metamorphic rocks. According to Searle
117 and Godin (2003) and Searle (2010) three main units (referred as three “*formations*” in Le
118 Fort, 1975) can be generally identified: (i) Unit 1 is made of kyanite-bearing metasediment
119 with subordinate quartzite, calcsilicate and marble. Migmatites are present in the upper part
120 of this unit; (ii) Unit 2 is mainly composed of medium to high-grade calcsilicate and minor
121 marble, (iii) Unit 3 consists mainly of orthogneiss and minor kyanite/sillimanite migmatite.
122 Structurally upwards (mainly within Unit 3) the GHS is intruded by Miocene leucogranite,
123 referred as Higher Himalayan Leucogranites (HHL, Le Fort, 1975; Visonà et al., 2012), forming
124 foliation-concordant sills and cross-cutting dykes up to large (kilometre sized) plutons.
125 Melting is mainly the result of muscovite and biotite dehydration melting (Patiño -Douce and
126 Harris, 1998; Visonà et al., 2012; Searle, 2013 and references therein). Moreover, Prince et al.
127 (2001) recognized also the presence of an older “water-fluxed melting episode” at higher
128 pressure (see also King et al., 2011).

129 The GHS evolution is often subdivided into two main metamorphic stages, classically (*e.g.*
130 Vannay and Hodges, 1996) referred as Eohimalayan HP-MT event (M1) of Eocene-Oligocene
131 age, in which the GHS reached the highest pressure (in the kyanite stability field) and the
132 Neohimalayan MP-HT Miocene event (M2). For instance, the GHS rocks studied by Liu et al.
133 (2007) experienced P-T conditions of 1.4 GPa and 750-800°C at 33 ± 2 Ma (zircon U-Pb
134 SHRIMP) and then the original kyanite-bearing paragenesis in metapelites (*e.g.*, Borghi et al.,
135 2003) was overprinted by sillimanite- and cordierite-bearing paragenesis, associated with
136 widespread melting, at 23 ± 2 Ma (Liu et al., 2007).

137 Deformation within the GHS is dominantly characterized by general shear (*e.g.* Larson and
138 Godin, 2009 and references therein). The GHS is tectonically overlaid by the Tethyan

139 Sedimentary Sequence (TSS) *via* a large scale ductile to brittle system of normal faults, named
140 South Tibetan Detachment System (Caby et al., 1983; Burg et al., 1984; Burchfiel et al., 1992;
141 Carosi et al., 1998, 2002; Searle, 2010) active in the same time span of the activity of the MCT.

142 4) The TSS consists of early Palaeozoic to late Mesozoic multi-phase folded
143 unmetamorphosed to low grade metamorphic sediments (Godin et al., 1999a,b; Antolín et al.,
144 2011; Dunkl et al., 2011) originally deposited on the northern passive margin of the Indian
145 Plate. Towards the N, the TSS is bounded by flysches and ophiolites (often with a blueschist
146 metamorphic imprint, Hodges, 2000) of the Indus-Tsangpo suture zone (Fig. 1).

147 **3. The GHS in the Kali Gandaki valley**

148 The N-S trending Kali Gandaki valley (Fig. 1b) cross-cuts the whole Himalayan units and
149 structures (Fig. 1a, b), offering a clear exposure of metamorphic rocks and their structural
150 relationships. For these reasons the valley is probably one of the most classic study area of
151 Himalayan geologists (*e.g.* Le Fort et al., 1986; Colchen et al., 1986; Vannay and Hodges, 1996;
152 Godin et al., 1999a,b; Godin et al., 2001; Godin, 2003; Larson and Godin, 2009; Searle, 2010;
153 Carosi et al., 2014a,b). Nevertheless up to now metamorphic P-T paths obtained with
154 pseudosection modelling are lacking and very few *in situ* mineral ages are available along this
155 transect.

156 In the Kali Gandaki valley the GHS appears as a homogeneous homoclinal slab with isoclinal
157 folds (Brown and Nazarchuk, 1993; Vannay and Hodges, 1996) reaching a structural thickness
158 of 10-15 km (*e.g.* Godin 2003). The main foliation (S_2) related to a second deformation phase
159 (D_2), strikes NW-SE and moderately dips to the NE, whereas the main mineral lineation (L_2)
160 strikes mostly E-W or SE-NE and plunges to the E (*e.g.* Vannay and Hodges 1996, Carosi et al.,
161 2014b). Relicts of an older deformative event D_1 , (Eohimalayan event of Vannay and Hodges,
162 1996) have been reported, for instance, by Vannay and Hodges, 1996. This event is testified

163 by sporadic evidences (such as high-angle mica and kyanite) of an older foliation (S_1) and as
164 an internal foliation in garnet porphyroblasts.

165 According to Vannay and Hodges (1996) the GHS in the Kali Gandaki could be subdivided into
166 four “formations” (Formation 1, 2, 3, corresponding to Unit 1, 2, 3 of Searle and Godin, 2003)
167 plus the upper Larjung Formation.

168 The lowest one (unit 1 of Fig. 1b) contains kyanite-bearing paragneiss and micaschist. The
169 second one (unit 2 of Fig. 1b), representing the thick core of the GHS in this transect, consists
170 of calcsilicate with the paragenesis of clinopyroxene, garnet, amphibole (often with titanite)
171 and minor metapelite.

172 The third unit (Unit 3, Fig. 1), made of orthogneiss and minor metapelite, is overlaid by *c.* 200
173 m of amphibole-bearing calcsilicate-gneiss (Larjung formation). According to Godin (2003)
174 the Larjung formation, interpreted by previous authors as the base of the TSS (*e.g.* Colchen et
175 al., 1986), is deformed together with the upper part of the GHS by the Annapurna Detachment
176 (AD), a 1500 m thick high-strain zone, representing a local segment of the STDS. Godin et al.
177 (2001) suggested an age of *c.* 22 Ma for the cessation of the ductile shearing along the AD.

178 Within Unit 3, a ductile shear zone with a top-to-the southwest sense of shear, named
179 Kalopani Shear Zone (KSZ in Fig. 1b), has been identified by Vannay and Hodges (1996) and is
180 interpreted as an out of sequence thrust (*e.g.* Vannay and Hodges, 1996; Godin et al., 1999).
181 Based on Ar-Ar white-mica geochronology an age older than 13-15 Ma has been suggested for
182 the shearing (Vannay and Hodges, 1996). New U-Th-Pb monazite ages indicate an Eocene age
183 (*c.* 30-40 Ma) for the shearing activity along the KSZ (Carosi et al., 2014b).

184 In the rocks belonging to the MCTZ–Lower GHS, Vannay and Hodges (1996) reported two
185 groups of samples based on their P-T record. Eohimalayan Group 1 with P-T conditions of
186 $610 \pm 40^\circ \text{C} - 0.94 \pm 0.09 \text{ GPa}$, is interpreted to have mainly equilibrated close to the

187 metamorphic peak (at the beginning of thrusting along the MCT), whereas the Neohimalayan
188 Group 2 ($540\pm 30^{\circ}\text{C}$ $0.65\pm 0.03\text{ GPa}$), which is more intensively sheared and retrogressed,
189 records the exhumation stage during or after the MCT activity. Based on Ar-Ar geochronology,
190 Vannay and Hodges (1996) suggested that the whole GHS rapidly cooled below the white-
191 mica closure temperature ($300\text{--}430^{\circ}\text{C}$ in Vannay and Hodges, 1996) during the Early–Middle
192 Miocene (13–15 Ma). Carosi et al. (2014a) identified “nanogranites” as inclusions in garnets
193 from kyanite-bearing rocks (bottom of the upper GHS) testifying that these rocks have
194 experienced melting starting at *c.* 41–36 Ma, based on U-Pb *in situ* monazite ages, further
195 supporting the idea that these rocks should be better classified as migmatitic paragneisses
196 (see also Searle, 2010).

197 **4. Outcrop description and sample petrography**

198 The studied gneiss comes from outcrops close to Titar Village (North of Dana village, Fig. 1b),
199 > 1 km north of the MCT as mapped by Colchen et al. (1986) and Vannay and Hodges (1996)
200 because top-to-the south kinematic indicators are present. Recently, Parsons et al. (2014)
201 assigned this thrust structure to the Chomrong Thrust (CT) shifting the MCT further to the
202 south.

203 In the outcrops (Fig. 2a–d) kyanite-garnet-biotite-white mica-bearing migmatitic metapelite
204 occur with thin intercalations (dm-thick) of garnet-amphibole-bearing gneiss.

205 Leucocratic layers (*in situ* leucosome) parallel to the main foliation commonly occur, which
206 form tight to isoclinal folds with axial planes parallel to the main foliation (Sp). These
207 leucosomes, mainly made of plagioclase and quartz, are stretched and the folds asymmetry
208 points a top-to-the S sense of shear. Centimetric garnet and kyanite are abundant (Fig. 2b,c),
209 reaching the largest grain size within the leucocratic layers. Mafic selvages (*e.g.* Sawyer,
210 2008) of biotite were also observed (Fig. 2a, b). Moreover, late leucocratic pods are present

211 (Fig. 2d). Several hand specimens, covering the range of structures observed at the mesoscale
212 were sampled (K28a to K28g).

213 Microscopically the main foliation is classified as a spaced anastomosing foliation (Passchier
214 and Trouw, 2005) and is defined by lepidoblastic levels of biotite and white mica, where
215 kyanite also occurs (Fig. 3a). Mineral lineation (Lp) is mainly defined by aligned kyanite and
216 stretched quartz and feldspar. The microscale kinematic indicators, such as shear bands (Fig.
217 3b) and sigma type porphyroclasts, support a top-to-the S sense of shear.

218 Garnet is subhedral and full of tiny inclusions, which are in some cases moderately iso-
219 oriented. Inclusions, often polyminerally, are made of plagioclase, rutile/ilmenite, white mica,
220 biotite and minor chlorite. Garnet rims are frequently replaced by biotite and plagioclase. Late
221 fractures, partially filled with chlorite/green biotite, are oriented at high angle with respect to
222 the main foliation. They often cause retrogression of inclusions, with ilmenite replacing rutile
223 and white mica being transformed to biotite/chlorite. The main foliation wraps around the
224 garnet porphyroblasts, supporting their pre-tectonic occurrence (with respect to the top-to-
225 the south shearing).

226 Kyanite forms large porphyroblasts (early synkinematic?) often aligned with the main
227 foliation. This mineral contains inclusions of paragonite, quartz, potassic white mica and
228 rutile. Plagioclase around quartz inclusions in kyanite was sporadically observed (*e.g.* Carosi
229 et al., 2014a). This feature suggests that also kyanite was present during melting (Carosi et al.,
230 2014a). Kyanite in the mesosome and less frequently in the leucosome is sometimes
231 deformed (kinked) and locally partially replaced by fine-grained white mica. Very late
232 sillimanite needles (Fig. 3c) rarely occur on kyanite, near the garnet and at the plagioclase
233 boundaries.

234 Quartz and plagioclase show microstructural evidence of dynamic recrystallisation. Quartz
235 has irregular, lobed grain boundaries interpreted as due to Grain Boundary Migration
236 recrystallisation (Passchier and Trouw, 2005). Undulose extinction, often squarish
237 (chessboard extinction, Fig. 3d), testifies a high temperature ($\geq 650^{\circ}\text{C}$) deformation regime
238 (Passchier and Trouw, 2005). Plagioclase shows lobed grain boundaries and in some cases
239 deformation twinning is present. Strain-free grains of both minerals were also observed.
240 Within the sheared leucosomes, euhedral/subhedral plagioclase crystals, with well-developed
241 crystal faces have been observed in some instances (Fig. 3f), testifying the heterogeneous
242 nature of the deformation.

243 The stable Ti-phase in the matrix is ilmenite. Rutile is present only as relict cores in the
244 former mineral and enclosed in kyanite and garnet. Other accessory phases are apatite (as the
245 major phosphate), monazite, pyrite, zircon and tourmaline. These minerals occur in both
246 garnet and matrix, whereas tiny xenotime grains were observed only in the matrix.

247 In spite of the strong deformation, several microstructural observations (see criteria review
248 of Holness et al., 2011 and references therein) suggest that melt was present in the samples:
249 (i) quartz grains with evidence of corrosion and rimmed by feldspar in quartzofeldspathic
250 domains and within kyanite (Carosi et al., 2014a); (ii) tiny films of feldspars with cusped low
251 dihedral angles; (iii) “string of beads” microstructures (Fig. 3e); (iv) euhedral feldspar grains;
252 and (v) “nanogranite” inclusions within peritectic garnet (Carosi et al., 2014a).

253 From a petrological point of view the inferred “peak mineral assemblage” is interpreted to
254 have been garnet-kyanite-biotite-plagioclase-quartz-white mica-rutile (+ melt), whereas
255 ilmenite and the rare sillimanite needles are considered as post-peak minerals which grew
256 during decompression and/or cooling. Chlorite locally on biotite and garnet suggests a very
257 late fluid infiltration along some foliation planes (*e.g.* Vannay and Hodges, 1996). In a single
258 case (sample K28c) hematite overgrowth on pyrite was detected within these alteration

259 zones. It is very important to stress the lack of K-feldspar in the studied rocks (neither
260 observed here nor reported by Vannay and Hodges, 1996 and Carosi et al., 2014a), while it
261 was observed, together with prismatic sillimanite, in migmatites in similar structural position
262 in the Sikkim Himalaya (Harris et al., 2004). The stromatic metatexite sample K28a was
263 selected for a detailed petrological and geochronological investigation.

264 **5. Methods**

265 *5.1 Mineral chemistry and compositional maps*

266 After a careful optical inspection, K28a thin sections were studied with a CAMECA SX100
267 electron microprobe (EMP) hosted at Institut für Mineralogie und Kristallchemie (Universität
268 Stuttgart) equipped with five wavelength dispersive spectrometers (WDS). The energy
269 dispersive spectrometer (EDS) of the EMP was used for qualitative identification of minerals.

270 Chemical compositional maps (X-Ray mapping) were acquired on selected minerals/areas
271 (micas and garnets) with a stepwise movement (100 ms per step) using an electron beam
272 with a beam current of 60 nA, 150 nA, 30 nA for garnet, monazite and micas, respectively, and
273 subsequent computer-aided evaluation. Garnets were mapped for Y, Ca, Mn, Fe, Mg, monazites
274 for Y, Th, U, Ce, Si, and micas for Ba, Na, Mg, Fe, Ti.

275 Quantitative chemical analyses of points and transects were acquired on minerals present in
276 all textural positions (matrix or included in porphyroblasts) using an acceleration voltage of
277 15 kV and a beam current of 15 nA. Monazite grains were analyzed following the procedure
278 described in Massonne et al. (2007). Synthetic and natural standards were used for EMP
279 calibration. The analytical uncertainties for the EMP measurements applied here are reported
280 by Massonne (2012). For analyzing Zr in rutile (see below) a beam current of 100 nA and an
281 acceleration voltage of 15 kV were selected. Structural formulae from mineral analyses were
282 calculated with the software CALCMIN (Brandelik, 2009).

283 5.2 *P-T* estimates

284 In order to derive a *P-T* path, phase equilibria constraints, using pseudosection modelling,
285 were derived for the selected sample (K28a). Previous authors have shown how powerful this
286 tool is for bracketing the *P-T* evolution of Himalayan migmatites (*e.g.* Harris et al., 2004;
287 Groppo et al., 2010, 2012; Guilmette et al., 2011).

288 *P-T* pseudosections were constructed for the *P-T* range of 0.3-1.3 GPa and 600-850°C, for a
289 fixed bulk composition with the software PERPLE_X (*e.g.* Connolly 2005, version from August
290 2011, downloaded from the web site <http://www.perplex.ethz.ch/>). For this purpose we used
291 the internally consistent thermodynamic database for minerals and water (CORK model,
292 Holland and Powell 1991) given by Holland and Powell (1998, and updates). The following
293 solid-solution (a-X) models were used: GLTsTsPg for amphibole, T for talc, Ctd(HP) for
294 chloritoid, TiBio(HP) for biotite, Chl(HP) for chlorite, hCrd for cordierite, Gt(HP) for garnet,
295 Opx(HP) for orthopyroxene, Omph(HP) for clinopyroxene, IlGkPy for ilmenite, Pheng(HP) for
296 potassic white mica (with a maximum paragonite content of 50% mol), and St(HP) for
297 staurolite (details on http://www.perplex.ethz.ch/perplex_solution_model_glossary.html).

298 The models used for feldspars (plagioclase and K-feldspar) and paragonitic mica were
299 reported by Massonne (2012 and references therein). Moreover, in order to calculate melting
300 relationships, the model melt(HP) for haplogranitic melt (White et al., 2001) was used. The
301 bulk composition was obtained with the X-ray fluorescence (XRF) spectrometer at the Earth
302 Science Department of Pisa University, using the procedure of Tamponi et al. (2002-2003).

303 Calculations were performed in the MnO-Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂
304 system (MnNCKFMASHT). Titanium was included in order to determine the *P-T* stability of Ti-
305 rich phases, while the O₂ (needed to consider ferric iron) was neglected because: *i*) magnetite
306 is absent; *ii*) the amount of ferric iron in minerals is low and *iii*) rutile+ilmenite (\pm pyrite)

307 should indicate low oxidation conditions (Diener and Powell, 2010; see also Groppo et al.,
308 2010). The XRF composition was somewhat simplified for fitting the ten-component model
309 chemical system as follows: *i)* CaO was reduced applying a correction for (ideally composed)
310 apatite; *ii)* various amounts of H₂O were considered in the pseudosections calculations. The
311 pseudosection results for 2.0 wt% of H₂O are presented here. Maybe such H₂O values are too
312 high, but, for instance, Braga and Massonne (2012) have demonstrated for the Ulten Zone in
313 the Eastern Alps that still higher water amounts can be present in HP anatectic metapelites.
314 Calculations with different water (*e.g.* 2.5 wt%) or oxygen contents (despite the
315 aforementioned negligence of O₂) have been explored resulting in minor changes of the
316 pseudosection topology (*e.g.* magnetite occurrence, small T shift of the solidus, see also
317 Massonne, 2014), with no significant changes in the considered phase-in boundaries (*e.g.*
318 cordierite-in; K-feldspar-in) unless too low water amounts are assumed, stabilizing K-feldspar
319 (not observed) in subsolidus assemblages (*e.g.* Massonne, 2014).

320 The pseudosections were contoured by molar amounts of phase components, like pyrope
321 component in garnet (isopleths thermobarometry, *e.g.* Gaidies et al., 2006) and by modal
322 amounts of phases (*e.g.* melt volume). Moreover, the “geothermobarometric potential of
323 anatectic melts” as described by Massonne et al. (2013) and applied by Cruciani et al. (2014)
324 has been explored comparing pseudosection predictions with results on “nanogranites”
325 remelting experiments by Carosi et al. (2014a). The latter authors fully re-homogenized the
326 crystallized melt inclusions during remelting experiments at a temperature of 820°C, a
327 pressure of 1.2 GPa, and 24h experimental run, obtaining a melt composition with Si/Al ratio
328 of 3.93 (± 0.25) and Na/K ratio of 2.7 (± 1.0) (Carosi et al., 2014a, their table 1).

329 The calculated raw P-T graphs were smoothed as shown by Connolly (2005). According to
330 Massonne (2013), uncertainties of 10% on the P- and 5% on the T-estimates were considered
331 for our P-T data resulting from the pseudosection modelling. For this reason, we also applied

332 an independent method related to the Zr-in-rutile thermometry (Zack et al., 2004) using the
333 pressure sensitive calibration of Tomkins et al. (2007). Recently, Ewing et al. (2013)
334 demonstrated that the Zr-in rutile thermometer has good chances to record “*peak*”
335 temperatures in high-grade rocks, especially for pristine (not-retrogressed/recrystallized)
336 rutile, while Hallett and Spear (2014) testified that this thermometer provides important
337 constraints for revealing the history of anatectic metapelites. Also Zr-in-rutile values, obtained
338 for sample K28c by Carosi et al. (2014a), are reported for comparison.

339 *5.3 Monazite in situ U-Th-Pb geochronology*

340 Monazite, (LREE, Th)PO₄, can be a reliable geochronometer (*e.g.* Parrish, 1990). In recent
341 years much attention has been paid in order to quantify its behaviour during deformation
342 events (*e.g.* Williams and Jercinovic, 2002, 2012; Dumond *et al.* 2008) and metamorphic
343 reactions (*e.g.* Foster and Parrish, 2003; Foster et al., 2000; Gibson et al., 2004; Spear and Pyle,
344 2002, 2010 and references therein). These studies have shown how monazite can record the
345 timing of metamorphic processes for a wide spectrum of metamorphic conditions of the
346 greenschist facies (*e.g.* Gasser et. al., 2012,) up to the granulite facies (*e.g.* Rubatto et al., 2001,
347 2013; Pyle and Spear 1999, 2003; Martins et al., 2009; Gasser et. al., 2012; Palin et al., 2014;
348 Alcock et al., 2013; Massonne, 2014). Moreover, many efforts have been undertaken in order
349 to link the chemistry of monazite to environmental parameters such as the temperature (see
350 Spear and Pyle, 2002, for a review) using for example the monazite-xenotime thermometer
351 (*e.g.* Spear and Pyle, 2002). *In situ* geochronological techniques offer the possibility of linking
352 U-Th-Pb isotopic ages to particular chemical and/or textural domains related to metamorphic
353 reactions and/or deformation events that could still be present also in unshielded matrix
354 grains (*e.g.* Langone et al., 2011). For these reasons, in order to put time constraints in the
355 evolution of the studied rock, monazite grains were carefully characterized.

356 Prior to isotopic dating of monazite grains, their textural position and internal features were
357 imaged with a scanning electron microscope hosted at Earth Department of Pisa University,
358 while the chemical characterization was achieved with the EMP as described above (section
359 5.1).

360 Monazite crystals were analysed *in situ* by laser-ablation, inductively coupled plasma mass
361 spectrometry (LA-ICPMS) directly on 30 µm thick thin sections at the CNR-Istituto di
362 Geoscienze e Georisorse U.O. Pavia (Italy) using an Ar-F 193-nm excimer laser (GeolLas 102
363 from Micro-Las) coupled with a magnetic sector ICP-MS (Element I from Thermo-Finnigan).
364 The full description of the analytical procedure is reported in Paquette and Tiepolo (2007)
365 and Tiepolo et al. (2003). Multiple laser spots are available for almost each grain using X-Ray
366 maps as guide for spot position. Moreover, where it was possible, the laser spot was located at
367 an area close to the EMP analytical spot.

368 Single analyses were performed by a one-minute acquisition of the background signal
369 followed by recording, for at least 30 seconds, the ablation signal of the masses related to the
370 isotopes ^{202}Hg , $^{204}(\text{Hg}+\text{Pb})$, ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th , and ^{238}U . The presence of common Pb
371 was evaluated in each analysis on the basis of the net signal of ^{204}Pb (i.e. subtracted for the
372 interference of ^{204}Hg and background). None of the sample revealed ^{204}Pb counts above the
373 background level. However, the relatively high Hg signal in the gas blank does not exclude the
374 effective presence of common Pb in the analysed monazite. Analytical conditions were 10 µm
375 diameter of spot size, 12 J cm⁻² of energy density, and 3 Hz of repetition rate. Time-resolved
376 signals were carefully inspected to verify the presence of perturbations related to inclusions,
377 fractures or mixing of different age domains. Laser-induced elemental fractionation and mass
378 bias were corrected using matrix-matched external monazite standard (Moacir monazite:
379 Seydoux- Guillaume et al., 2002a,b) considering the values, re-calibrated for isotopic
380 disequilibrium, reported by Gasquet et al. (2010). In the analytical run eight to nine spots of

the external standard were analysed. Only those close to the reference values (at least 4 in each run) were considered in order to reduce errors related to the standard reproducibility (Table A.1). External standards and unknowns were integrated over the same time intervals to ensure the efficient correction of fractionation effects. Data reduction was carried out with the GLITTER® software (van Achterbergh et al., 2001). In order to better estimate the uncertainty affecting the $^{206}\text{Pb}/^{238}\text{U}$, $^{207}\text{Pb}/^{235}\text{U}$ and $^{208}\text{Pb}/^{232}\text{Th}$ isotope ratios, the external reproducibility of the standard was propagated relative to individual uncertainties for the isotope ratios. After this error propagation each analysis is accurate within the quoted errors. The Isoplot 3.0 software by Ludwig (2003) has been used for age calculation and graphic representation.

6. Results

6.1 Mineral compositions (except monazite)

White mica has a composition showing a slight chemical variability, which deviates from ideal muscovite (Fig. 4a,c-Fig.5a). The contents of Si are between 3.10 and 3.16 a.p.f.u. (Fig. 5a) with the highest values observed mainly in the inner part of large grains (Fig. 4a,c) where typically relatively high Ti contents (up to 0.06 a.p.f.u.) were observed (Fig.4c). Mg/(Mg+Fe) (hereafter Mg#) is between 0.57-0.64. Moreover, paragonite (Fig. 5a) with XNa (*i.e.* Na/(Na+K) values between 0.87 and 0.93 was detected within kyanite grains.

Biotite shows Mg# between 0.51 and 0.56 (Fig. 5b) with the highest value found in biotite included in garnet. Ti (a.p.f.u.) values range between 0.11 and 0.17 (Fig. 5b) with the lowest values occurring in biotite enclosed in garnet.

Garnet composition is slightly variable as shown by X-ray maps (Fig. 4c,d and Fig. 5c). This mineral is rich in almandine component ($X_{\text{Alm}} > 0.72$). The garnet core is characterized by 23 mol% of pyrope component, 3 mol% of grossular (+ andradite) component and 2 mol% of

405 spessartine component. Towards the rim, contents of Mn slightly increase (3% mol of
406 spessartine), pyrope contents decrease down to 17 mol% and Ca contents are almost
407 constant.

408 The composition of *plagioclase* varies with XAb between 0.86 and 0.92. The orthoclase
409 component is always below 1% (Fig. 5d). A clear core to rim zoning is observed in matrix
410 plagioclase where rims are somewhat richer in anorthite component.

411 *Ilmenite* contains some pyrophanite component (1.5-4 mol%). *Rutile* shows Cr (205-718 ppm)
412 and Nb (750-15050 ppm) contents typical of metapelitic rocks (*e.g.* Meinhold 2010 and
413 references therein). However, some analyses are also very rich in Nb (> 4000 ppm). Zr values
414 between 270 – 650 ppm were determined. No clear correlation has been found between
415 chemistry and rutile position within garnet, whereas lower Zr contents are present both in
416 “pristine” rutile and more typically in the ilmenite-rimmed rutile. These chemical
417 observations are valid for both K28a and K28c samples.

418

419 6.2 *P-T constraints and P-T path*

420 6.2.1 *P-T Pseudosection*

421 The P-T pseudosection for K28a (Fig. 6) shows a dominance of four-variance fields with
422 quartz and plagioclase being ubiquitous. Garnet is absent at low P (fields n° 7, 8, and 9 in Fig.
423 6). Cordierite appears towards the high T side at low P, whereas biotite is completely
424 consumed above *c.* 825°C. Melt is predicted to appear around 650°C. Rutile is stable above
425 0.70-0.80 GPa. Subsolidus kyanite occurs in a restricted P-T range (600-650°C, *c.* 0.7-0.8 GPa)
426 whereas it is largely present at suprasolidus conditions. Interestingly, paragonite is present in
427 the HP-LT corner of the pseudosection.

428 The assumed peak assemblage (garnet-kyanite-biotite-plagioclase-quartz-white mica-rutile +
429 melt) is represented by a quatrovariant field (labeled as LWmPlGrtBtKyQzRt in Fig. 6) in the
430 P-T range of *c.* 0.8-1.3 GPa and *c.* 650-800 °C, delimited by the disappearance of kyanite
431 towards higher P and lower T (field n° 1) and by the appearance of K-feldspar (and ilmenite)
432 towards higher high T and lower P. Melting along a nearly isobaric path, in this field is
433 accompanied by the consumption of biotite-quartz-plagioclase and the formation of garnet,
434 kyanite and white mica (for reactions of this type see Palin et al., 2014 and also King et al.,
435 2011). Ilmenite and rare sillimanite, interpreted as post peak phases, formed later at
436 relatively low P conditions (Fig. 6).

437 The relevant isopleths are displayed in Figs. 7a-d. XCa isopleths in garnet show a moderately
438 positive dP/dT slope (Fig. 7a) and increasing Ca contents in garnet with rising pressure.
439 Typically, the pyrope (XMg) content in garnet increases with rising temperature and pressure
440 (Fig. 7a). This trend is opposed to the core to rim zoning of the studied garnets (Fig. 4d). For
441 this reason, it is likely that the use of garnet compositions for deriving a P-T path, results only
442 in minimum P-T conditions along a retrograde segment of the path.

443 Isopleths for Si contents in potassic white mica (Fig. 7b) show relatively flat dP/dT slopes at
444 subsolidus and suprasolidus conditions. The albite content in plagioclase is predicted to
445 increase with rising pressure (Fig 7b). The molar Si/Al ratio of the melt is mainly controlled
446 by pressure (Fig 7.c), whereas the isopleths for Na/K ratio of the melt show moderate dP/dT
447 slopes in agreement with results by Massonne et al. (2013) and Cruciani et al. (2014). The
448 calculated melt volume (Fig. 7c) is below 10% in volume before crossing the white mica-out
449 curve. After crossing a significant increase of the melt is observed ("*effective solidus*" of White
450 et al., 2001).

451 6.2.2 Zr-in-Rutile

452 Consistent temperatures in the range of *c.* 650-720°C were obtained (Fig. 8), despite a large
453 variation of Nb in rutile, a possible effect of which on the Zr-in-rutile thermometer was not
454 experimentally constrained so far. Calculated temperatures close to 720°C can be easily
455 explained to represent the “metamorphic T peak”, consistent with the pseudosection results
456 (see below, Fig. 9). Lower temperatures (*c.* 650°C) were obtained for both rutile enclosed in
457 “pristine” garnet and occurring in the matrix. This is interpreted as “prograde” T recorded in
458 pristine rutile, whereas Zr remobilization (*e.g.* Luvinzotto and Zack, 2009) during rutile
459 retrogression could be invoked for post “peak” temperatures obtained from rutile of the
460 matrix.

461 6.2.3 *P-T path*

462 On the basis of the aforementioned petrography and the chemical compositions of relevant
463 phases a P-T path was reconstructed (Fig. 9). The P-T conditions for garnet core formation are
464 close to 700 °C and 1.0 GPa. Slightly higher pressures or lower temperatures are indicated by
465 the highest Si contents in white mica, whereas somewhat higher temperature and pressure
466 conditions are indicated by the Si/Al and Na/K ratios of the melt (despite the larger scatter,
467 see above), the anorthite content in the plagioclase cores and the Zr-in-rutile thermometer. In
468 this way a conservative P-T estimate for peak conditions reached by the studied sample K28a,
469 is *c.* 710-720°C and 1.0-1.1 GPa (Fig. 9), where nearly 7 vol% melt is predicted to occur.
470 Similar P-T results of 1.14 GPa -722°C (± 0.21 GPa, $\pm 34^\circ\text{C}$, sigfit 1.6) have been also obtained
471 with AvePT thermobarometry (Powell and Holland, 1994) using the highest XMg, lowest XMn
472 garnet composition with the average of matrix biotite, plagioclase and white mica. The
473 compositions of the outermost garnet rim and white mica with the lowest Si contents on the
474 main foliation provide a constraint that the retrograde path passed through P-T conditions of
475 650-670°C and 0.7-0.8 GPa, near the sillimanite-kyanite transition curve (Fig. 9). During this
476 segment of the path characterized by decompression and cooling, garnet, white mica and the

477 melt were consumed, whereas a minor increase in melt volume ($\leq 1\%$ in vol) is predicted for
478 the first stage (turning point) of the decompression. Plagioclase zoning, “reverse” zoning in
479 garnet, and chemical compositions of potassic white mica and biotite are compatible with this
480 retrograde segment of the P-T path. The proposed P-T path, lacking of large isothermal
481 decompression, is also in accordance with the lack of K-feldspar in the sample (see also
482 experimental results of Patiño-Douce and Harris, 1998).

483 Although speculative, the prograde P-T path characterized by both increasing P and T (dashed
484 bold red curve in Fig. 9) is reconstructable based on the inclusion of rutile and paragonite in
485 kyanite, the presence of “nanogranite” inclusions in peritectic garnet, and temperatures
486 recorded by “pristine” rutiles (see paragraph 6.2.2).

487 *6.3 Monazite dating*

488 *6.3.1 Monazite textural position and chemistry*

489 Monazite was found in different microstructural positions (Fig. 10). This mineral is very
490 common in both phyllosilicate and quartz-feldspar rich domains of the matrix forming grains
491 up to 200 μm in diameter, which often have equilibrium grain faces with adjacent grains.
492 Monazite shows inclusions of biotite, quartz and white mica. Small inclusions of monazite
493 within porphyroblasts are much less common. In few cases (Mnz 19 in Fig. c) a grain
494 completely armored in kyanite was detected. In one case a monazite grain was found to be
495 located at the kyanite-quartz interface (Mnz 6). Monazite inclusions in garnet are also present.
496 However, the complete shielding from the matrix was not always certain due to fractures in
497 garnet. In one case (Mnz 1, Fig. 10a) monazite was found as inclusion in rutile within a garnet
498 porphyroblast.

499 Backscattered electron (BSE) images show clear zoning in most grains with domains of
500 different graytones (Fig. 10). X-ray compositional maps (Fig. 10a-f) and EMP analyses (Table

2, Fig. 11) reveal that such zoning is controlled by different distribution of HREE+Y vs LREE (Fig. 11) and Th. Matrix monazite often shows resorbed cores (*e.g.* Fig. 10c, e) with intermediate (Fig. 11) values of HREE+Y (1.5-1.9 wt% Y₂O₃) and variable Th contents. A brighter intermediate zone (mantle) is characterized by very low values of HREE+Y (Fig. 10e, Fig. 11) (0.2-0.7, wt% Y₂O₃) and generally higher Th contents. A darker outer zone, often forming a discontinuous rim on the intermediate zone, shows relatively high values of HREE+Y (2.5-3.3 wt% Y₂O₃). Monazite included in garnet is chemically similar to the cores of matrix monazite and monazite enclosed in kyanite. Mnz 19, fully included in kyanite, shows both the resorbed intermediate HREE+Y core and the mantle zone of the matrix monazite, whereas the high HREE rim is lacking. Such a rim domain has been observed in grain Mnz 6, which is only partially included in kyanite and in a few monazite grains within fracture zones in garnet.

In summary, based on the previously described textural and chemical arguments monazite shows three main growth domains/generations (Fig. 11) *i.e.*: (i) Mnz I with intermediate Y₂O₃ contents, is included in garnet and kyanite and occurs as resorbed cores in matrix monazite; (ii) Mnz II with very low HREE+Y contents is present in the matrix as mantle on resorbed cores and in monazite hosted in kyanite; (iii) Mnz III, forming discontinuous high HREE+Y rims, occurs only in the matrix. All three domains/generations are not always present in a single matrix grain. For instance, the Mnz I domain is absent in some grains. Nevertheless, monazite with all domains occurs in several matrix “Rosetta grains” (*sensu* Dahl et al., 2005) such as grain “Mnz 21” (Fig. 10e). Since the monazite HREE+Y budget in metapelites is largely controlled by garnet crystallization (*e.g.* Forster and Parrish, 2003; Spear and Pyle, 2010), the observed HREE+Y variations could be linked to several steps of garnet growth/resorption.

6.3.2 *In-situ U-Th-Pb geochronology and age interpretations*

525 The measured isotopic data are reported in Table 3, whereas results for the Moacir standard
526 are reported in Table A.1. A total of 22 monazite grains covering the whole textural-chemical
527 variability, were selected for *in situ* dating obtaining data from a total of 45 spots.
528 According to Foster et al. (2000) the obtained results are plotted in a ^{238}U - ^{206}Pb vs ^{232}Th - ^{238}Pb
529 concordia diagram (Fig. 12a). Also a probability density plot of Th-Pb ages is given (Fig. 12b).
530 These ages are preferred for the discussion due to the ^{230}Th effect on U-Pb ages (*e.g.* Schärer
531 1984), even if this should be rather minor (*e.g.* Kellett et al., 2010). As it can be seen from Fig.
532 12, a large spread of ages ranging from 43 Ma to 18 Ma was determined. Ages in the range of
533 43-36 Ma were obtained from resorbed cores of matrix monazite and monazite included in
534 garnet and kyanite (Mnz I). Ages around 29 Ma are recorded by the mantle of matrix monazite
535 as well as monazite grains included within kyanite (Mnz II). Younger ages in the range of 25
536 Ma - 18 Ma are related to the high HREE+Y rims (Mnz III) of matrix grains. Large age
537 variations are very common in medium-high grade metamorphic monazite (*e.g.* Foster et al.,
538 2000, 2002; Martins et al., 2009; Rubatto et al., 2013; Massonne, 2014; Palin et al., 2014).
539 Different mechanism could be invoked to explain such spread (see discussion in Foster et al.,
540 2002): (i) presence of different chemical-age domains, as the result of monazite continuous or
541 discontinuous growth; (ii) mixing of different age domains during laser ablation; (iii) Pb loss
542 due to diffusive processes in a grain. Several studies (*e.g.* Spear and Pyle, 2002; Seydoux-
543 Guillaume et al., 2002a; Cherniak et al., 2004; Gardés et al., 2007 and references therein) have
544 shown that Pb diffusion in monazite is very slow and comparable with zircon. Thus, process
545 (iii) can be excluded to explain the observed age spread. Mixing of different domains during
546 ablation is possible despite careful spot position location and signal inspection, considering
547 the relatively large ablation volume. However, common observations of systematic
548 correlation of intra-crystalline zoning (for example in Y-HREE) with ages (*e.g.* Foster et al.,
549 2000, 2002; Gibson et al., 2004; Williams and Jercinovic, 2002, 2012;) make the mechanism

(ii) largely unlikely for being the main reason of the age spread. The observations in the present case, that the three different texturally and chemically (with different HREE+Y and Th contents) recognized populations (Mnz I-Mnz II-Mnz III see above) correspond to “age” populations, strongly support the idea that monazite records several growth stages along the experienced P-T path.

With these considerations, we interpret the obtained monazite ages as follow: Mnz I, due to its HREE+Y contents, grew during the prograde pre-melting (upper greenschist?-middle amphibolite facies?) part of the P-T path and experienced partial resorption during anatexis leading to the production of peraluminous melt (*e.g.* Pyle and Spear, 2003; Spear and Pyle, 2010). Mnz II, with its low HREE+Y contents, could be linked to the growth from a melt where garnet was part of the peritectic assemblage (*e.g.* Martins et al., 2009; Gasser et al., 2012) and partially shielded from the growth of the last generation (Mnz III) due to the entrapment in porphyroblasts. If this is true, crustal melting occurred in the time span between 36 and 28 Ma. Although Mnz III gave occasionally old ages as high as 30 Ma (mixing of domains ?), it is systematically younger than monazite of the other two groups. Because of high HREE+ Y contents this monazite (Mnz III) could be linked to the garnet breakdown during the exhumation stage accompanied by release of Y_2O_3 (including melt consuming back-reactions and/or a likely fluid infiltration) promoting monazite (Mnz III) re-growth. Assuming equilibrium conditions (see criteria reported by Spear and Pyle, 2002) between matrix xenotime and Mnz III with a X_{HREE+Y} of 0.096-0.087 (Table 2, Fig. 11, mole fractions calculated according to Pyle et al. 2001), the temperature span for crystallisation of Mnz III could be estimated in the range of *c.* 700-600°C (Table 2). This range is somewhat lower than the derived “peak” temperature condition and can be referred to cooling.

573

574 7. Discussions

575 *7.1 P-T-D-t path of kyanite-bearing migmatite in the Kali Gandaki*

576 Sheared migmatitic paragneiss of Unit 1 in the Kali Gandaki, according to petrological
577 observations and thermodynamic calculations on sample K28a (see paragraph 6.2),
578 experienced partial melting at “near peak” conditions around 710-720°C and 1.0-1.1 GPa
579 (Figs. 9, 10) where kyanite-bearing tonalitic leucosomes developed. In addition to the
580 presence of kyanite, the occurrence of rutile (within the supra-solidus porphyroblasts) is also
581 diagnostic for melting at high pressure conditions, since the rutile-ilmenite transition is a
582 useful pressure monitor in Barrovian pelitic rocks under reducing conditions (Weller et al.,
583 2013; Massonne, 2014). According to Weller et al. (2013) rutile is typical of HP Barrovian
584 metamorphism that is necessarily linked to a rock sequence metamorphosed below an
585 overthrusting continental plate (see discussion in Weller et al., 2013).

586 The here estimated “near peak P-T” conditions (710-720°C, 1.0-1.1 GPa) are somewhat higher
587 (610°C – 0.9 GPa, see paragraph 3) than the ones previously proposed by Vannay and Hodges
588 (1996) using classical thermobarometry. In addition, the estimated temperatures are also *c.*
589 100°C lower than those resulting from remelting experiments of “nanogranites” reported by
590 Carosi et al. (2014a). Although these authors considered the obtained melt composition as the
591 true product of Himalayan anatexis, they regarded the obtained re-melting temperature
592 unlikely for kinetic reasons linked to the experimental procedure (*e.g.* experimental time
593 duration). This consideration is confirmed here, since temperatures of 800°C and more are
594 not in agreement with the presence of white mica and the absence of K-feldspar in the rock
595 and the result of the Zr-in-rutile thermometry.

596 The studied migmatitic paragneiss, after melting at near peak conditions, experienced
597 decompression and cooling, associated with pervasive heterogeneous shearing, to reach P-T
598 conditions of 650-670°C and 0.7-0.8 GPa. These conditions are consistent with the chessboard

599 extinction observed in quartz. The retrograde part of the path is also compatible with an
600 ilmenite-rutile transition at 0.7-0.8 GPa above 650°C (Massonne, 2014). Moreover, according
601 to the melt isomodes (Fig. 7d) little melt (≤ 1.0 vol%) can be expected to have formed during
602 the early stages of the here-inferred decompressional segment of the P-T path (see also
603 Groppo et al., 2010, 2012 for higher T samples). It is worthy of note that the inferred P-T path
604 has a similar shape as the P-T path proposed by Guillot (1999) (see his figure 5).

605 The timing of the P-T history has been constrained by *in situ* U-Th-Pb monazite dating.
606 According to the obtained monazite ages the near peak-pressure melting of the sheared
607 migmatitic paragneiss occurred in depths of 35-40 km during the ongoing prograde
608 metamorphism (with a minimum starting age of 43 Ma, see also Carosi et al., 2010) in the time
609 span between 36 and 28 Ma. This geochronological estimate, despite the different approaches
610 (in situ dating vs mineral separation) and dating systematics, is compatible with the ages of c.
611 35-32 Ma reported by Godin et al. (2001) and of c. 41–36 Ma by Carosi et al. (2014a) for the
612 GHS melting at HP conditions in the Kali Gandaki. Moreover, the present data potentially
613 indicate that anatexis conditions could have been sustained for c. 8-10 Myr (see also Palin et
614 al., 2014). Starting from nearly 25 Ma up to 18 Ma the investigated rock experienced a
615 substantial change of the dP/dT slope of their path and started to be exhumed
616 (decompression and cooling segment of the P-T path) accompanied by melt crystallization,
617 garnet breakdown and Mnz III formation (Fig. 10).

618 7.2 Formation of kyanite-bearing migmatites in the Himalaya

619 Along the Himalayan belt several kyanite (+ late sillimanite) bearing migmatites, occurring at
620 different structural positions within the GHS, have been studied so far with respect to their
621 metamorphic and temporal evolution. They are given here according to their regional position
622 from east to west and compared with our results:

623 Guilmette et al. (2011) studied kyanite-bearing anatectic paragneiss from the Eastern
624 Himalayan Syntaxis. These HP granulitic rocks experienced “peak conditions” of 820°C at
625 pressures higher than 1.4 GPa (possibly 1.5 – 1.6 GPa), followed by decompression, cooling
626 and melt solidification at 810°C and 0.9 GPa. Geochronological information on HP melting in
627 the Eastern Himalayan Syntaxis was reported by Palin et al. (2014). These authors identified
628 two high-grade metamorphic (and melting) events in their studied migmatites. The first one
629 (71-50 Ma), within the sillimanite stability field, is related to the pre-collisional tectonic
630 history of the Lhasa block. The second event, producing migmatites in the kyanite stability
631 field and tonalitic leucosomes, occurred at minimum P-T conditions of nearly 700°C and 1.04
632 GPa (based on THERMOCALC Average P-T method) in the time span of 44–33 Ma (based on U-
633 Th-Pb monazite geochronology). This event, that lasted nearly 10 Ma, was related by Palin et
634 al. (2014) to the India-Asia collision.

635 Davidson et al. (1997) and Daniel et al. (2003) reported similarly deformed anatectic
636 metasediments from the Bhutan Himalaya just few hundreds of metres above the MCT and,
637 thus, in a structural position comparable with the studied samples. Davidson et al. (1997)
638 suggested minimum P-T conditions of 0.8 GPa and >700°C, while Daniel et al. (2003), based
639 on conventional geothermobarometry (see their Fig. 9), proposed P-T conditions of c. 750-
640 800°C and 1.2 - 1.3 GPa, at which melting (*i.e.* in the kyanite stability field) occurred. The latter
641 authors reported also U-Pb ages on separated monazite and xenotime grains suggesting the
642 formation of kyanite-bearing migmatites around 18-16 Ma.

643 In the kyanite-sillimanite migmatitic gneisses of the Sikkim Himalaya (Harris et al., 2004) pre-
644 decompressional garnet growth is dated at 23 ± 3 Ma (Sm-Nd systematics), while garnet
645 growth at near-peak temperatures (750 °C - 0.8 GPa) occurred at 16 ± 2 Ma during the
646 melting stage. According to Harris et al. (2004) the major melting stage is interpreted as the
647 result of the decompression to the sillimanite stability field. Despite their similar approach for

P-T estimates (*i.e.* pseudosection), in the Kali Gandaki migmatites studied here, petrographic and petrological data testify that melting already occurred within the kyanite stability field (as also suggested by Daniel et al., 2003; see also Groppo et al., 2010), well before the Miocene STDS-related decompression and the sillimanite melting stage (overprinting ?).

In Far Eastern Nepal along the Arun-Makalu transect Groppo et al. (2010) described the P-T-t evolution of a kyanite-bearing migmatitic sample, coming from the structurally highest portion of the lower GHS, where also K-feldspar was part of the main peak assemblage. Their studied sample reached P-T conditions of 820°C, 1.3 GPa around 31 Ma (Early Oligocene) and then followed decompression, cooling and melt back-reactions down to c. 800 °C, 1.0 GPa in the time span of 27-29 Ma (see their figure 9). Also, Imayama et al. (2012) studied kyanite – sillimanite migmatites from Far-Eastern Nepal. These authors, combining pseudosections and trace-element constrained U-Pb zircon ages, demonstrated that their kyanite-sillimanite migmatites experienced melting at c. 21-18 Ma and isothermal decompression from P-T peak of 0.8–1.4 GPa and 720–770°C.

Sillimanite (after kyanite)-bearing migmatites from the Kharta valley in Tibet near Mt. Everest, studied by Liu et al. (2007), had experienced somewhat higher pressures and temperatures (1.4 GPa and 750-800°C) than our migmatites. However, the temporal evolution is well comparable with our results as the HP event for the Kharta valley rocks were dated at 33 ± 2 Ma (Liu et al., 2007). Also the retrograde evolution of these rocks at 23 ± 2 Ma (Liu et al., 2007) coincides with the time range of 18-25 Ma determined for our rocks.

The aforementioned different P-T-t conditions estimated for the formation of kyanite-bearing GHS migmatites, could be partially explained by different geothermobarometric and dating methods applied by the various authors. In part these differences can be due to samples taken from different structural positions. Despite these differences, the present data indicate that (1) melting at HP conditions produced tonalitic or leucogranitic melts (as function of the P-T

673 conditions) already during the early stages of the India-Asia collision forming the Himalaya
674 and, (2) other portions of the belt probably reached HP condition (and melting) at different
675 times (up to the Miocene).

676 *7.3 Tectonic implications*

677 Several discontinuities identified within the GHS (*e.g.* Carosi et al., 2007; 2010; Corrie and
678 Kohn, 2011; Imayama et al., 2012; Larson et al., 2013; Montomoli et al., 2013; see Montomoli
679 et al., 2014 for a review) have shown that the GHS has a much more complex crustal
680 architecture compared to simple models where a single coherent tectonic unit is bounded by
681 only two tectonic discontinuities with opposite sense of shear. These recent findings are also
682 compatible with diachronic melting within the GHS (Kohn et al., 2005; Corrie and Kohn, 2011;
683 Imayama et al., 2012; Rubatto et al., 2013). As anticipated above our samples are localized ~ 1
684 km northern than the MCT location according to Colchen et al. (1986) and Vannay and Hodges
685 (1996), recently mapped as CT by Parsons et al. (2014) whereas the MCT has been shifted
686 dozen km to the South according to Searle (2010). The different localization of the MCT could
687 arise some ambiguities on the tectonic meaning of our present results. Anyway, Montomoli et
688 al. (2014) discussed how the location of MCT vs P-T-D-t discontinuities could be problematic
689 since several processes (*e.g.* shear zone widening; ductile thinning due to pure shear
690 component of deformation) could complicate the structural pattern in the GHS. They
691 suggested that to characterize tectonic discontinuities field observations are not unique and a
692 multidisciplinary approach (joining structural, metamorphic and chronological information)
693 could help to solve ambiguity. Indeed, to better unravel the structural (and melting) evolution
694 of the GHS, monazite geochronology could help in assessing the ages of the activity of the
695 different tectonic discontinuities (*e.g.* Kohn et al., 2005; Corrie and Kohn, 2011). In particular,
696 in the present case study, the rim monazite ages (*c.* 25-18 Ma, Mnz III), interpreted as
697 retrograde, are useful for this assessment. The here presented Mnz III ages are older than the

698 quoted ages for the MCT activity in the Kali Gandaki (*c.* 21 –16 Ma in Gibson et al. 2014; *c.* 22
699 Ma in Godin et al., 2006) and share much more similarities with intra GHS in sequence
700 contractional shear zones, like the High Himalayan Discontinuity (HHD) of Montomoli et al.
701 (2013, 2014). The kinematics, the P-T-t path and the age of the studied sample testify, for the
702 first time, the occurrence of the HHD in the Kali Gandaki valley (Fig. 13). This occurrence at
703 25-18 Ma, in a structural higher position and older with respect to the MCT, dated at 22-16 Ma
704 along the same section, proves a southward shifting of the shearing and exhumation withing
705 the GHS (Fig. 13).

706 It is also important to stress that in this contribution the monazite ages related to the starting
707 of the kyanite growth are significantly older compared to other kyanite-bearing gneiss ages
708 within the MCT zone reported along other Himalayan transects (*e.g.* *c.* 22-15 Ma in Larson et
709 al., 2013; 17-13Ma in Montomoli et al., 2013). This means that kyanite in the MCT zone grew
710 at different times.

711 It also appears that the GHS melting history is not so simple as proposed in the current
712 geodynamic models (*e.g.* extrusion and channel flow). Improved future models should fully
713 account for the different melting scenarios and melting timing affecting the whole GHS such
714 as: high pressure – decompressional – and low pressure (with peritectic andalusite) melting
715 (see Visonà et al., 2012 and references therein). With respect to this point, we note that
716 melting was produced at different times within the different GHS slices. The associated
717 decrease of rock viscosity (*e.g.* Jamieson et al., 2011 and references therein) could be
718 responsible for strain softening processes with the consequence of localization of deformation
719 resulting in shear-zone nucleation and decoupling of the diverse GHS slices.

720 **Conclusions**

721 A comprehensive P-T-t-D path for kyanite-bearing migmatitic paragneiss of the GHS in the
722 Kali Gandaki has been reconstructed integrating meso- and micro-structural and petrographic
723 data, P-T estimates and *in situ* monazite geochronology (Fig. 13). The gneiss underwent
724 prograde metamorphism from (at least) 43 to 28 Ma and experienced partial melting at P-T
725 conditions of 710-720°C/1.0-1.1 GPa in the time span of 36-28 Ma, producing kyanite-bearing,
726 K-feldspar-poor leucosomes (Fig. 13). This rock was subject to decompression and cooling
727 associated with pervasive shearing during 25-18 Ma prior and at an upper structural position
728 with respect to the MCT activity (Fig. 13). This study testifies for the first time the occurrence
729 of a structural and tectonic discontinuity within the GHS in the Kali Gandaki valley and
730 confirms its regional extent (HHD; Montomoli et al., 2013, 2014) and the occurrence of
731 diachronous exhumation of the two portions of the GHS divided by the HHD. Exhumation in
732 the GHS was not triggered only by the MCT but started, before it was formed, in the upper
733 part of the GHS.

734 It is also suggested that the migmatitic paragneiss experienced a clockwise P-T loop as shown
735 in Figure 9 represents a potential source of “high-Ca” melts recognized in some parts of the
736 Himalaya (*e.g.* King et al., 2011), but the extent and the chronology of this HP melting event
737 along the Himalaya strike deserves further attention in the future.

738

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745 **Bibliography**

746 Antolín, B., Appel, A., Montomoli, C., Dunkl, I., Ding, L., Gloaguen, R., El Bay, R., 2011. Kinematic
747 evolution of the eastern Tethyan Himalaya: constraints from magnetic fabric and structural
748 properties of the Triassic flysch in SE Tibet. In: Poblet, J., Lisle, R. (Eds.), Kinematic Evolution
749 and Structural Styles of Fold-and-Thrust Belts: Geological Society of London Special
750 Publications 349, pp. 99–121. <http://dx.doi.org/10.1144/SP349.6>.

751 Alcock, J., Muller, P.D., Jercinovic, M.J., 2013. Monazite ages and pressure–temperature–time
752 paths from anatectites in the southern Ruby Range, Montana, USA: evidence for delamination,
753 ultramafic magmatism, and rapid uplift at ca. 1780 Ma. Canadian Journal of Earth Sciences 50,
754 1069-1084, 10.1139/cjes-2013-0035

755 Beaumont, C., Jamieson, R.A., Nguyen, M.H., Lee, B., 2001. Himalayan tectonics explained by
756 extrusion of a low-viscosity crustal channel coupled to focused surface denudation. Nature
757 414, 738–742.

758 Borghi, A., Castelli, D., Lombardo, B., Visona, D., 2003. Thermal and baric evolution of garnet
759 granulites from the Kharta region of S Tibet, E Himalaya. European Journal of Mineralogy 15,
760 401–418.

761 Braga, R., Massonne, H.-J., 2012. H₂O content of deep-seated orogenic continental crust: the
762 Ulten Zone, Italian Alps. International Geology Review 54, 633-641.

763 Brandelik, A., 2009. CALCMIN—an EXCEL™ Visual Basic application for calculating mineral
764 structural formulae from electron microprobe analyses source. Computers and Geosciences
765 35, 1540-1551.

766 Brown, R.L., Nazarchuk, J.H., 1993. Annapurna detachment fault in the Greater Himalaya of
767 Central Nepal. Geological Society of London 74, 41-473.

768 Burchfiel, B.C., Zhiliang, C., Hodges, K.V., Yuping, L., Royden, L.H., Changron, D., Jiene, X., 1992.
 769 The South Tibetan Detachment System, Himalayan Orogen. Geological Society of America
 770 Special Paper 269, 41 pp.

771 Burg, J.P., Brunel, M., Gapais, D., Chen, G.M., Liu, G.H., 1984. Deformation of leucogranites of the
 772 crystalline Main Central Sheet in southern Tibet (China). *Journal of Structural Geology* 6, 535–
 773 542.

774 Caby, R., Pêcher, A., Le Fort, P., 1983. Le grand chevauchement central himalayen: Nouvelles
 775 données sur le métamorphisme inverse à la base de la Dalle du Tibet. *Revue de Géologie*
 776 *Dynamique et de Géographie Physique* 24, 89–100.

777 Carosi, R., Lombardo, B., Molli, G., Musumeci, G., Pertusati, P.C., 1998. The south Tibetan
 778 detachment system in the Rongbuk valley, Everest Region. Deformation features and
 779 geological implication. *Journal of Asian Earth Sciences* 16, 299–311.

780 Carosi, R., Montomoli, C., Visonà, D., 2002. Is there any detachment in the Lower Dolpo
 781 (western Nepal)? *Comptes Rendus Geoscience* 334, 933–940.

782 Carosi, R., Montomoli, C., Visonà, D., 2007. A structural transect in the Lower Dolpo: Insights
 783 on the tectonic evolution of Western Nepal. *Journal of Asian Earth Sciences* 29, 407–423,
 784 doi:10.1016/j.jseaes.2006.05.001.

785 Carosi, R., Montomoli, C., Rubatto, D., Visonà D., 2010. Late Oligocene high-temperature shear
 786 zones in the core of the Higher Himalayan Crystallines (Lower Dolpo, western Nepal).
 787 *Tectonics* 29, DOI: 10.1029/2008TC002400.

788 Carosi, R., Montomoli, C., Langone, A., Turina, A., Cesare, B., Iaccarino, S., Fascioli, L., Visonà, D.,
 789 Ronchi, A., Rai, S.M., 2014a. Eocene partial melting recorded in peritectic garnets from

790 kyanite-gneiss, Greater Himalayan Sequence, central Nepal. Geological Society, London,
 791 Special Publications, 412, doi:10.1144/SP412.1

792 Carosi, R., Gemignani, L., Godin, L., Iaccarino, S., Larson, K., Montomoli, C., Rai, S.M., 2014b. A
 793 geological journey in the deepest canyon of the earth: the Kali Gandaki section. In: (Eds.)
 794 Chiara Montomoli, Rodolfo Carosi, Rick Law, Sandeep Singh, and Santa Man Rai, Geological
 795 field trips in the Himalaya, Karakoram and Tibet, Journal of the Virtual Explorer, Electronic
 796 Edition, ISSN 1441-8142, volume 47, paper 9.

797 Cherniak, D.J., Watson, E.B., Grove, M., 2004. Pb diffusion in monazite: a combined RBS/SIMS
 798 study. *Geochimica et Cosmochimica Acta* 68, 829–840.
 799 <http://dx.doi.org/10.1016/j.gca.2003.07.012>.

800 Colchen, M., Le Fort, P., Pêcher, A., 1986. Notice explicative de la carte géologique Annapurna–
 801 Manaslu–Ganesh (Himalaya du Népal) au 1:200 000e (bilingual: French–English), CNRS, Paris,
 802 1986.

803 Connolly, J.A.D., 2005. Computation of phase equilibria by linear programming: A tool for
 804 geodynamic modeling and its application to subduction zone decarbonation. *Earth and*
 805 *Planetary Science Letters* 236, 524–541.

806 Corrie, S.L., Kohn, M.J., 2011. Metamorphic history of the central Himalaya, Annapurna region,
 807 Nepal and implications for tectonic models. *Geological Society of America Bulletin* 123,
 808 18563-1879.

809 Cruciani, G., Fancello, D., Franceschelli, M., Scodina M., Spano, M.E., 2014.
 810 Geothermobarometry of Al-silicate-bearing migmatites from the Variscan chain of NE
 811 Sardinia, Italy: a P-T pseudosection approach. *Periodico di Mineralogia* 83, 19-40.

812 Dahl, S.P., Terry, M.P., Jercinovic, M.J., Williams, M.L., Hamilton, M.A., Foland, K.A., Clement, S.-
813 M., Friberg, L.M., 2005. Electron probe (Ultrachron) microchronometry of metamorphic
814 monazite: Unraveling the timing of polyphase thermotectonism in the easternmost Wyoming
815 Craton (Black Hills, South Dakota). *American Mineralogist* 90, 1712–1728.

816 Daniel, C.G., Hollister, L.S., Parrish, R.R., Grujic, D., 2003. Exhumation of the Main Central
817 Thrust from lower crustal depths, eastern Bhutan Himalaya. *Journal of Metamorphic Geology*
818 21, 317–334.

819 Davidson, C., Grujic, D.E., Hollister, L.S., Schmid, S.M., 1997. Metamorphic reactions related to
820 decompression and synkinematic intrusion of leucogranite, High Himalayan Crystallines,
821 Bhutan. *Journal of Metamorphic Geology* 15, 593–612.

822 Diener, J.F.A., Powell, R., 2010. The influence of ferric iron on the stability of mineral
823 assemblages. *Journal of Metamorphic Geology* 28, 599–613.

824 Dumond, G., McLean, N., Williams, M. L., Jercinovic, M.J., Bowring, S.A., 2008. High resolution
825 dating of granite petrogenesis and deformation in a lower crustal shear zone: Athabasca
826 granulite terrane, western Canadian Shield. *Chemical Geology* 254, 175–196.

827 Dunkl, I., Antolín, B., Wemmer, K., Rantitsch, G., Kienast, M., Montomoli, C., Ding, L., Carosi, R.,
828 Appel, E., El Bay, R., Xu, Q., von Eynatten, H., 2011. Metamorphic evolution of the Tethyan
829 Himalayan flysch in SE Tibet. In: Gloaguen, R., Ratschbacher, L. (Eds.), *Growth and Collapse of*
830 *the Tibetan Plateau: Geological Society of London Special Publications* 353, 45–69.
831 <http://dx.doi.org/10.1144/SP353.4>.

832 Ewing, T., Hermann, J., Rubatto, D., 2013. The robustness of the Zr-in-rutile and Ti-in-zircon
833 thermometers during high-temperature metamorphism (Ivrea-Verbano Zone, northern Italy).
834 *Contributions to Mineralogy and Petrology* 165, 757–779.

835 Foster, G., Parrish, R.R., 2003. Metamorphic monazite and generation of P–T–t paths. In:
836 Vance, D., Müller, W. & Villa, I. M. (eds) *Geochronology: Linking the Isotopic Record with*
837 *Petrology and Textures*. Geological Society, London, Special Publications 220, 25–47.

838 Foster, G., Kinny, P., Vance, D., Prince, C., Harris, N., 2000. The significance of monazite U–Th–
839 Pb age data in metamorphic assemblages; a combined study of monazite and garnet
840 chronometry. *Earth and Planetary Science Letters* 181, 327–340.

841 Foster, G.L., Gibson, H.G., Parrish, R.R., Horstwood, M.S.A., Fraser, J., Tindle, A., 2002. Textural,
842 chemical and isotopic insights into the nature and behavior of metamorphic monazite.
843 *Chemical Geology* 191, 183–207. [http://dx.doi.org/10.1016/S0009-2541\(02\)00156-0](http://dx.doi.org/10.1016/S0009-2541(02)00156-0).

844 Gaidies, F., Abart, R., De Capitani, C., Schuster, R., Connolly, J.A.D., Reusser, E., 2006.
845 Characterization of polymetamorphism in the Austroalpine basement east of the Tauern
846 Window using garnet isopleth thermobarometry. *Journal of Metamorphic Geology* 24, 451–
847 475.

848 Gardés, E., Montel, J.M., Seydoux-Guillaume, A.M., Wirth, R., 2007. Pb diffusion in monazite:
849 New constraints from the experimental study of $\text{Pb}^{2+} \leftrightarrow \text{Ca}^{2+}$ interdiffusion. *Geochimica et*
850 *Cosmochimica Acta* 71, 4036–4043. DOI: 10.1016/j.gca.2007.06.036

851 Gasser, D., Bruand, E., Rubatto, D., Stüwe, K., 2012. The behaviour of monazite from
852 greenschist facies phyllites to anatectic gneisses: An example from the Chugach Metamorphic
853 Complex, southern Alaska. *Lithos* 134–135, 108–122.

854 Gansser, A., 1964. *Geology of Himalayas*. Wiley Interscience, London, 289.

855 Gasquet, D., Bertrand, J.-M., Paquette, J.-L., Lehmann, J., Ratzov, G., De Ascensão Guedes, R.,
856 Tiepolo, M., Boullier, A.-M., Scaillet, S., Nomade, S., 2010. Miocene to Messinian deformation
857 and hydrothermal activity in a pre-Alpine basement massif of the French western Alps: new

858 U-Th-Pb and argon ages from Lauzière massif. Bulletin de la Société Géologique de France
 859 181, 227–241.

860 Gibson, H.D., Carr, D.S., Brown, L.R., Hamilton, A.M., 2004. Correlations between chemical and
 861 age domains in monazite, and metamorphic reactions involving major pelitic phases: an
 862 integration of ID-TIMS and SHRIMP geochronology with Y-Th-U X-ray mapping. Chemical
 863 Geology 211, 237–260.

864 Gibson, R., Godin, L., Cottle, M.J., Kellett, D.A., 2014. Along-strike strain variation and timing of
 865 deformation in the lower Himalayan metamorphic core, west-central Nepal. In Montomoli C.,
 866 et al., eds., proceedings for the 29th Himalaya-Karakoram-Tibet Workshop, Lucca, Italy.
 867 Journal of Himalayan Earth Sciences (Special Volume), 45-47.

868 Godin, L., Brown, R.L., Hanmer, S., Parrish, R., 1999a. Back folds in the core of the Himalayan
 869 orogen: An alternative interpretation. Geology 27, 151-154.

870 Godin, L., Brown, R.L., Hanmer, S., 1999b. High strain zone in the hanging wall of the
 871 Annapurna detachment, central Nepal Himalaya, in Macfarlane, A., Sorkhabi, R. B., and Quade,
 872 J., eds., Himalaya and Tibet: Mountain roots to mountain tops: Geological Society of America
 873 (Special Paper) 328, 199–210.

874 Godin, L., Parish, R.R., Brown, L., Hodges, K.V., 2001. Crustal thickening leading to exhumation
 875 of the Himalayan Metamorphic core of central Nepal: insight from U-Pb geochronology and
 876 $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology. Tectonics 20, 729-747.

877 Godin, L., 2003. Structural evolution of the Tethyan sedimentary sequence in the Annapurna
 878 area, central Nepal Himalaya. Journal of Asian Earth Sciences 22, 307–328.

879 Gratz, R., Heinrich, W., 1997. Monazite-xenotime thermometry: experimental calibration of
 880 the miscibility gap in the system $\text{CePO}_4\text{--YPO}_4$. American Mineralogist 82, 772–780.

881 Groppo, C., Rubatto, D., Rolfo, F., Lombardo, B., 2010. Early Oligocene partial melting in Main
882 Central Thrust Zone (Arun valley, eastern Nepal Himalaya). *Lithos* 118, 287–301.

883 Groppo, C., Rolfo, F., Indares, A., 2012. Partial melting in the Higher Himalayan Crystallines of
884 Eastern Nepal: the effect of decompression and implications for the 'Channel Flow' model.
885 *Journal of Petrology* 53, 1057–1088.

886 Guillot, S., 1999. An overview of the metamorphic evolution in Central Nepal. *Journal of Asian*
887 *Earth Sciences* 17, 713–725.

888 Guilmette, C., Indares, A., Hébert, R., 2011. High-pressure anatectic paragneisses from the
889 Namche Barwa, Eastern Himalayan Syntaxis: Textural evidence for partial melting, phase
890 equilibria modeling and tectonic implications. *Lithos* 124, 66–81.

891 Hallett, B.W., Spear, F.S., 2014. The P-T history of anatectic pelites of the Northern East
892 Humboldt Range, Nevada: Evidence for tectonic loading, decompression, and anatexis. *Journal*
893 *of Petrology* 55, 3–36.

894 Harris, N.B.W., Caddick, M., Kosler, J., Goswami, S., Vance, D., Tindle, A.G., 2004. The pressure-
895 temperature-time path of migmatites from the Sikkim Himalaya. *Journal of Metamorphic*
896 *Geology* 22, 249–264.

897 Heim, A., Gansser, A., 1939. Central Himalaya: geological observations of the Swiss expedition
898 1936. *Memoir Society Helvetica Science Nature* 73, 1–245.

899 Herron, M.M., 1988. Geochemical classification of terrigenous sands and shales from core or
900 log data. *Journal of Sedimentary Research* 58, 820–829, doi:10.1306/212F8E77-2B24-11D7-
901 8648000102C1865D.

902 Hodges, K.V., 2000. Tectonics of Himalaya and southern Tibet from two perspectives.
903 *Geological Society of America Bulletin* 112, 324–350.

904 Holness, M., Cesare, B., Sawyer, E.W., 2011. Melted rocks under the microscope:
905 microstructures and their interpretation. *Elements* 7, 247–252.

906 Holland, T.J.B., Powell, R., 1991. A compensated-Redlich-Kwong (CORK) equation for volumes
907 and fugacities of CO₂ and H₂O in the range 1 bar to 50 kbar and 100-1600°C. *Contributions to*
908 *Mineralogy and Petrology* 109, 265-273.

909 Holland, T.J.B., Powell, R., 1998. An internally consistent thermodynamic data set for phases of
910 petrological interest. *Journal of Metamorphic Geology* 16, 309-343.

911 Jamieson, R.A., Beaumont, C., Medvedev, S., Nguyen, M.H., 2004. Crustal channel flows: 2.
912 Numerical models with implications for metamorphism in the Himalayan-Tibetan orogen.
913 *Journal of Geophysical Research* 109, B06407, doi:10.1029/2003JB002811.

914 Jamieson, R.A., Unsworth, M.J., Harris, N.B.W., Rosenberg, C.L., Schulmann, K., 2011. Crustal
915 melting and the flow of mountains. *Elements* 7, 253-260.

916 Kellett, D.A., Grujic, D., Warren, C., Cottle, J., Jamieson, R., Tenzin, T., 2010. Metamorphic
917 history of a syn-convergent orogen-parallel detachment: the South Tibetan detachment
918 system, Bhutan Himalaya. *Journal of Metamorphic Geology* 28, 785–808.
919 <http://dx.doi.org/10.1111/j.1525-1314.2010.00893.x>.

920 King, J., Harris, N., Argles, T., Parrish, R., Zhang, H. 2011. Contribution of crustal anatexis to the
921 tectonic evolution of Indian crust beneath southern Tibet. *Geological Society of America*
922 *Bulletin* 123, 218–239.

923 Kohn, M.J., Wieland, M., Parkinson, C.D., Upreti, B.N., 2005. Five generation of monazite in
924 Langtang gneisses: implication for chronology of the Himalayan metamorphic core. *Journal of*
925 *Metamorphic Geology* 23, 399–406.

926 Imayama, T., Takeshite, T., Yi, K., Cho, D.-L., Kitajima, K., Tsutsumi, Y., Kayama, M., Nishido, H.,
927 Okumura, T., Yagi, K., Itaya, T., Sano, Y., 2012. Two-stage partial melting and contrasting
928 cooling history within the Higher Himalayan Crystalline Sequence in the far-eastern Nepal
929 Himalaya. *Lithos* 134–135, 1–22.

930 Langone, A., Braga, R., Massonne, H.-J., Tiepolo, M., 2011. Preservation of old (prograde
931 metamorphic) U–Th–Pb ages in unshielded monazite from high-pressure paragneiss of the
932 Variscan Ulten Zone. *Lithos* 127, 68–85.

933 Larson, K.P., Godin, L., 2009. Kinematics of the Greater Himalayan sequence, Dhaulagiri Himal:
934 implications for the structural framework of central Nepal. *Journal of the Geological Society*
935 166, 25–43.

936 Larson, K.P., Gervais, F., Kellett, D.A., 2013. A P–T–t–D discontinuity in east-central Nepal:
937 implications for the evolution of the Himalayan mid-crust. *Lithos* 179, 275–292.

938 Le Fort, P., 1975. Himalaya: the collided range. *American Journal of Science* 275, 1–44.

939 Le Fort, P., Pêcher, A., Upreti, B.N., 1986. A section through the Tibetan Slab in central Nepal
940 (Kali Gandaki Valley): mineral chemistry and thermobarometry of the Main Central Thrust
941 Zone. *Sciences de la Terre, Nancy*, 47, 211–228.

942 Liu, Y., Siebel, W., Massonne, H.-J., Xiao, X., 2007. Geochronological and petrological
943 constraints for the tectonic evolution of the central Greater Himalayan Sequence in the Kharta
944 area, southern Tibet. *Journal of Geology* 115, 215–230.

945 Ludwig, K.R., 2003. Isoplot/Ex version 3.0. A geochronological toolkit for Microsoft Excel.
946 Berkeley Geochronology Center, Special Publications, 4.

947 Luvinzotto, G.L., Zack, T., 2009. Nb and Zr behavior in rutile during high-grade metamorphism
948 and retrogression: An example from the Ivrea–Verbano Zone. *Chemical Geology* 261, 303–317.

949 Martins, L., Vlach, S.R.F., Janasi, V.A. 2009. Reaction microtextures of monazite: correlation
950 between chemical and age domains in the Nazaré Paulista migmatite, SE Brazil. *Chemical*
951 *Geology* 261, 271–285.

952 Massonne, H.-J., 2012. Formation of amphibole and clinozoisite-epidote in eclogite owing to
953 fluid infiltration during exhumation in a subduction channel. *Journal of Petrology* 53, 1969-
954 1998, doi: 10.1093/petrology/egs040.

955 Massonne, H.-J., 2013. Constructing the pressure-temperature path of ultrahigh-pressure
956 rocks. *Elements* 9, 267–279.

957 Massonne, H.-J., 2014. Wealth of P–T–t information in medium-high grade metapelites:
958 Example from the Jubrique Unit of the Betic Cordillera, S Spain. *Lithos* 208–209, 137–157.

959 Massonne, H.-J., Cruciani, G., Franceschelli, M., 2013. Geothermobarometry on anatectic melts
960 – a high pressure Variscan migmatite from northeast Sardinia. *International Geology Review*
961 55, 1490-1505.

962 Massonne, H.-J., Kennedy, A., Nasdala, L., Theye, T., 2007. Dating of zircon and monazite from
963 diamondiferous quartzofeldspathic rocks of the Saxonian Erzgebirge — hints at burial and
964 exhumation velocities. *Mineralogical Magazine* 71, 407–425.

965 Meinhold, G., 2010. Rutile and its applications in earth sciences. *Earth Science Reviews* 102, 1-
966 28.

967 Montomoli, C., Iaccarino, S., Carosi, R., Langone, A., Visonà, D., 2013. Tectonometamorphic
968 discontinuities within the Greater Himalayan Sequence in Western Nepal (Central Himalaya):
969 insights on the exhumation of crystalline rocks. *Tectonophysics* 608, 1349–1370.

970 Montomoli, C., Carosi, R., Iaccarino, S., 2014. Tectonometamorphic discontinuities in the
 971 Greater Himalayan Sequence: a local or a regional feature? Geological Society, London, Special
 972 Publications, 412, <http://dx.doi.org/10.1144/SP412.3>

973 Najman, Y., Apple, E., Boudagher-Fadel, M., Bown, P., Carter, A., Garzanti, E., Godin, L., Han, J.,
 974 Liebke, U., Oliver, G., Parrish, R., Vezzoli, G., 2010. Timing of India-Asia collision: Geological,
 975 biostratigraphic, and palaeomagnetic constraints. *Journal of Geophysical Research: Solid*
 976 *Earth* 115, DOI: 10.1029/2010JB007673

977 Palin, R.M., Searle, M.P., St-Onge, M.R., Waters, D.J., Roberts, N.M.W., Horstwood, M.S.A.,
 978 Parrish, R.R., Weller, O.M., Chen, S., Yang, J., 2014. Monazite geochronology and petrology of
 979 kyanite- and sillimanite-grade migmatites from the northwestern flank of the eastern
 980 Himalayan syntaxis. *Gondwana Research* 26, 323–347.

981 Paquette, J.L., Tiepolo, M., 2007. High resolution (5 μm) U–Th–Pb isotope dating of monazite
 982 with excimer laser ablation (ELA)–ICPMS. *Chemical Geology* 240, 222–237.

983 Parrish, R.R., 1990. U–Pb dating of monazite and its applications to geological problems.
 984 *Canadian Journal of Earth Sciences* 27, 1431–1450.

985 Parsons, A.J., Law, R.D., Searle, M.P., Phillips, R.J., Lloyd, G.E., 2014. Geology of the Dhaulagiri-
 986 Annapurna-Manaslu Himalaya, Western Region, Nepal, 1:200,000. *Journal of Maps*, DOI:
 987 10.1080/17445647.2014.984784.

988 Passchier, C.W., Trouw, R.A.J., 2005. Microtectonics. In: Springer (Ed.), p. 366.

989 Patiño-Douce, A., Harris, N., 1998. Experimental constraints on Himalayan anatexis. *Journal of*
 990 *Petrology* 39, 689–710.

991 Powell, R., Holland, T.J.B., 1994. Optimal geothermometry and geobarometry. *American*
 992 *Mineralogist* 79, 120-133.

993 Prince, C.P., Harris, N., Vance, D., 2001. Fluid enhanced melting during prograde
 994 metamorphism. *Journal of the Geological Society London* 158, 233–241.

995 Pyle, J.M., Spear, F.S., 1999. Yttrium zoning in garnet: coupling of major and accessory phase
 996 during metamorphic reactions. *Geologica Material Research* 1, 1–49.

997 Pyle, J.M., Spear, F.S., Rudnick, R.L., McDonough, F.W., 2001. Monazite-xenotime-garnet
 998 equilibrium in metapelites and a new monazite-garnet thermometer. *Journal of Petrology* 42,
 999 2083-2107.

1000 Pyle, J.M., Spear, F.S., 2003. Four generations of accessory-phase growth in lowpressure
 1001 migmatites from SW New Hampshire. *American Mineralogist* 88, 338–351.

1002 Rosenberg, C.L., Handy, M.R., 2005. Experimental deformation of partially melted granite
 1003 revisited: Implications for the continental crust. *Journal of Metamorphic Geology*, 23, 19-28.

1004 Rubatto, D., Williams, I.S., Buick, I.S., 2001. Zircon and monazite response to prograde
 1005 metamorphism in the Reynolds Range, central Australia. *Contributions to Mineralogy and*
 1006 *Petrology* 140, 458–468.

1007 Rubatto, D., Chakraborty, S., Dasgupta, S., 2013. Timescale of crustal melting in the Higher
 1008 Himalayan Crystallines (Sikkim, Eastern Himalaya) inferred from trace element-constrained
 1009 monazite and zircon chronology. *Contributions to Mineralogy and Petrology* 165, 349–372.
 1010 <http://dx.doi.org/10.1007/s00410-012-0812-y>.

1011 Sawyer, E.W., 2008. Atlas of migmatites. *The Canadian Mineralogist*. In: Special Publication,
 1012 vol. 9. NRC Research Press, Ottawa, Ontario, p. 371.

1013 Schärer, U., 1984. The effect of initial ^{230}Th disequilibrium on young U–Pb ages: the Makalu
 1014 case, Himalaya. *Earth and Planetary Science Letters* 67, 191–204.

1015 Searle, M.P., 2010. Low-angle normal faults in the compressional Himalayan orogen; Evidence
1016 from the Annapurna–Dhaulagiri Himalaya, Nepal. *Geosphere* 6, 296–315.

1017 Searle, M.P., 2013. Crustal melting, ductile flow, and deformation in mountain belts: Cause and
1018 effect relationships. *Lithosphere* 5, 547–554, doi:10.1130/RF.L006.1

1019 Searle, M.P., Godin, L., 2003. The South Tibetan Detachment System and the Manaslu
1020 leucogranite: a structural reinterpretation and restoration of the Annapurna–Manaslu
1021 Himalaya, Nepal. *Journal of Geology* 111, 505–523.

1022 Searle, M.P., Law, R.D., Godin, L., Larson, K.P., Streule, M.J., Cottle, J.M., Jessup, M.J., 2008.
1023 Defining the Himalayan Main Central Thrust in Nepal. *Journal of the Geological Society*,
1024 London 165, 523–534.

1025 Seydoux-Guillaume, A.M., Paquette, J.L., Wiedenbeck, M., Montel, J.M., Heinrich, W., 2002a.
1026 Experimental resetting of the U–Th–Pb system in monazite. *Chemical Geology* 191, 165–181.

1027 Seydoux-Guillaume, A.M., Wirth, R., Nasdala, L., Gottschalk, M., Montel, J.M., Heinrich, W.,
1028 2002b. XRD, TEM and Raman study of experimental annealing of natural monazite. *Physics
1029 and Chemistry of Minerals* 29, 240–253.

1030 Seydoux-Guillaume, A.M., Wirth, R., Heinrich, W., Montel, J.M., 2002c. Experimental
1031 determination of thorium partitioning between monazite and xenotime using analytical
1032 electron microscopy and X-ray diffraction Rietveld analysis. *European Journal of Mineralogy*
1033 14, 869– 878.

1034 Spear, F.S., Pyle, J.M., 2002. Apatite, monazite, and xenotime in metamorphic rocks. *Reviews in
1035 Mineralogy and Geochemistry* 48, 293–335.

1036 Spear, F.S., Pyle, J.M., 2010. Theoretical modeling of monazite growth in a low-Ca metapelite.
1037 *Chemical Geology* 273, 111–119.

1038 Stephenson, B.J., Searle, M. P., Waters, D.J., Rex, D.C., 2001. Structure of Main Central Thrust
 1039 zone and extrusion of the High Himalayan deep crustal wedge, Kishwar-Zaskar Himalaya.
 1040 *Journal of Geological Society, London* 158, 637-652.

1041 Tamponi, M., Bertoli, M., Innocenti, F., Leoni, L., 2002-2003. X-Ray fluorescence analysis of
 1042 major elements in silicate rocks using fused glass discs. *Atti Società Toscana di Scienze*
 1043 *Naturali, Memorie Serie A* 108, 73-79.

1044 Tiepolo, M., Bottazzi, P., Palenzona, M., Vannucci, R., 2003. A laser probe coupled with ICP –
 1045 Double-Focusing Sector-Field mass spectrometer from in situ analysis of geological samples
 1046 and U–Pb dating of zircon. *The Canadian Mineralogist* 41, 259–272.

1047 Tomkins, H.S., Powell, R., Ellis, J., 2007. The pressure dependence of the zirconium-in-rutile
 1048 thermometer. *Journal of Metamorphic Geology* 25, 703-713.

1049 van Achterbergh, E., Ryan, C.G., Jackson, S.E., Griffin, W., 2001. Data reduction software for LA-
 1050 ICPMS. In: Sylvester, P. (ed.) *Laser ablation ICPMS in the Earth sciences: Principles and*
 1051 *Applications*. Mineralogical Association of Canada Short Course Series 29, 239–243.

1052 Vance, D., Mahar, E., 1998. Pressure-temperature paths from P-T pseudosections and zoned
 1053 garnets: potential, pitfalls and examples from Zaskar Himalaya, NW India. *Contributions to*
 1054 *Mineralogy and Petrology* 132, 225-245.

1055 Vannay, J.C., Hodges, K.V., 1996. Tectonometamorphic evolution of the Himalayan
 1056 metamorphic core between the Annapurna and Dhaulagiri, Central Nepal. *Journal of*
 1057 *Metamorphic Geology* 14, 635–656.

1058 Visonà, D., Carosi, R., Montomoli, C., Peruzzo, L., Tiepolo, M., 2012. Miocene andalusite
 1059 leucogranite in central-east Himalaya (Everest–Masang Kang area): low-pressure melting
 1060 during heating. *Lithos* 144, 194–208.

1061 Weinberg, R., Hasalová, P., 2015. Water-fluxed melting of the continental crust: A review.
1062 Lithos 212–215, 158–188.

1063 Weller, O.M., St-Onge, M.R., Waters, D.J., Rayner, N., Searle, M.P., Chung, S.-L., Palin, R.M., Lee,
1064 Y.-H., Xu, X., 2013. Quantifying Barrovian metamorphism in the Danba Structural Culmination
1065 of eastern Tibet. *Journal of Metamorphic Geology* 31, 909–935.

1066 White, R.W., Powell, R., Holland, T.J.B., 2001. Calculation of partial melting equilibria in the
1067 system Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O (NCKFMASH). *Journal of Metamorphic*
1068 *Geology* 19, 139–153.

1069 Williams, M.L., Jercinovic, M.J., 2002. Microprobe monazite geochronology: putting absolute
1070 time into microstructural analysis. *Journal of Structural Geology* 24, 1013–1028.

1071 Williams, M.L., Jercinovic, M.J., 2012. Tectonic interpretation of metamorphic tectonites:
1072 integrating compositional mapping, microstructural analysis and in situ monazite dating.
1073 *Journal of Metamorphic Geology* 30, 739–752.

1074 Yakymchuk, C., Brown, M., 2013. Consequences of open system melting in tectonics. *Journal of*
1075 *the Geological Society of London* 171, 21–40.

1076 Yin, A., 2006. Cenozoic tectonic evolution of the Himalayan orogen as constrained by along-
1077 strike variation of structural geometry, exhumation history, and foreland sedimentation.
1078 *Earth-Science Reviews* 76, 1–131.

1079 Zack, T., Moraes, R., Kronz, A., 2004. Temperature dependence of Zr in rutile: empirical
1080 calibration of a rutile thermometer. *Contributions to Mineralogy and Petrology* 148, 471–488.

1081 **Figures and tables captions**

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1083 Table 1: Representative silicate analyses (in wt%) for sample K28a. Mineral structural
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1085 = 3 O; n.a. = not analyzed.

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1091 Table 3: LA-ICP-MS isotopic results and monazite ages. Textural position of the grain is
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1094 Figure 1: a) Himalaya geological map with the location of the study area indicated (after
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1102 metapelite and calcsilicate (unit 3, GHS); 4 = calc-silicate and marble (unit 2, GHS); 5 =
1103 kyanite-garnet gneiss (unit 1, GHS); 6 = quartzite of LHS in Vannay and Hodges, (1996) or
1104 lower GHS in Parsons et al. (2014); 7 = alluvial debris; 8 = STDS; 9 = minor normal fault; 10 =
1105 Kalopani shear zone (KSZ); 11 = MCT in Vannay and Hodges, (1996) or CT in Parsons et al.
1106 (2014); 12 = main foliation; 13 = object lineation; 14 = location of study samples.

1107 Figure 2: Characteristics of the studied rocks. a) Outcrop view of the leucosome aligned along
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1115 partially replacing kyanite (sample K28g, plane-polarized light); d) Quartz chessboard
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1117 crossed polars); e) Quartz string of pearls at a feldspar-feldspar boundary (sample K28a,
1118 crossed polars); f) Euhedral plagioclase faces within the leucosome (sample K28g, crossed
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1120 plagioclase, Qz = quartz, Sil = sillimanite, Wm = white mica.

1121 Figure 4: X ray compositional maps for white mica (a for Ti, c for Mg) and garnet (b for Mn; d
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1123 Figure 5: Main silicate chemistry: a) white mica and paragonite; b) biotite; c) garnet; d)
1124 plagioclase.

1125 Figure 6: P-T pseudosection for sample K28a. The bulk composition (in wt%) used for the
1126 modelling is reported in the upper part of the P-T graph. According to Herron (1988) the
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1128 feldspar, Ilm = ilmenite, L = melt, Opx = orthopyroxene, Pg = paragonite, Rt = rutile, St =
1129 staurolite.

1130 Figure 7: Compositional and modal isopleths in the pseudosection of Fig. 6. a) XMg (red line)
1131 and XCa (green line) in garnet; b) Si atoms per formula units (a.p.f.u.) in potassic white mica
1132 (black line) and XAn in plagioclase (dashed black line); c) Si/Al (purple line) and Na/K (blue
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1134 (dashed red line).

1135 Figure 8: Zr-in-rutile temperature (T in °C, using the Tomkins et al., 2007 calibration at 11
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1137 to the calculated temperature (see discussion in Tomkins et al., 2007).

1138 Figure 9: P-T path based on isopleths of Fig. 7 and Zr-in rutile thermometry (see paragraph
1139 6.2). Melt-in curve refers to the first appearance of melt according to the Perple_X
1140 calculations.

1141 Figure 10: Textural position, BSE images and Th, Y chemical maps of selected monazite. a)
1142 Mnz 1 in rutile included in garnet; b) Mnz 3 and Mnz 4 in garnet; c) Mnz 19 completely
1143 shielded in kyanite; note the absence of “high Y” rims; d) Mnz 7 in biotite; note the inclusion of
1144 biotite in monazite; e) Mnz 21 in the matrix; note the discontinuous high-Y rim; f) matrix grain
1145 Mnz 16. Th-Pb ages are reported and the quoted errors refer to a 2σ confidence level.

1146 Figure 11: Monazite chemical variation of heavy rare earth elements (HREE) *versus* light rare
1147 earth elements (LREE). Core-mantle-rim “appellative” refers only to a geometric spot location
1148 of EMP analyses.

1149 Figure 12: a) ^{206}Pb - $^{238}\text{U}/^{208}\text{Pb}$ - ^{232}Th concordia diagram; b) Probability density plot for Th-Pb
1150 ages. In the box inside, monazite populations (Mnz I, Mnz II, Mnz III) are indicated.

1151 Figure 13: P-T-t-D path of the GHS kyanite-bearing migmatites (block diagrams modified after
1152 Vannay and Hodges, 1996) based on pseudosection modelling, Zr-in-rutile temperatures and
1153 monazite geochronological data (see text).

1154 **Appendix**

1155 Table A.1: Isotopic LA-ICP-MS results for the Moacir standard.

1156